# Coupling of Volatile Transport and Internal Heatflow on Triton

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## Abstract

Recently Brown *et al.* (*Science* 250, 1991 ) showed that Triton's internal heat source could amount to 5-20% of the absorbed insolation on Triton, thus significantly affecting volatile transport and atmospheric pressure. Subsequently, Kirk and Brown (Proc. 22cd LPSC, 1991 ) used simple analytical models of the effect of internal heat on the distribution of volatiles on Triton's surface, confirming the speculation of Brown *et al.* that Triton's internal heatflow could strongly couple to the surface volatile distribution. To further explore this idea, we present numerical models of the *permanent* distribution of nitrogen ice on Triton that include the effects of sunlight, the two-dimensional distribution of internal heatflow, the coupling of internal heatflow to the surface distribution of nitrogen ice, and the finite viscosity of nitrogen ice. From these models we conclude that: (1) The strong vertical thermal gradient induced in Triton's polar caps by internal heatflow facilitates viscous spreading to lower latitudes, thus opposing the poleward transport of volatiles by sunlight, and, for plausible viscosities and nitrogen inventories, producing quasi-permanent caps having considerable latitudinal extent; (2) It is probable that there is a strong coupling between the surface distribution of nitrogen ice on Triton and internal heatflow; (3) Asymmetries in the spatial distribution of Triton's heatflow, possibly driven by large-scale, volcanic activity or convection in Triton's interior, can result in permanent polar caps of unequal latitudinal extent, including the case of only one permanent polar cap; (4) In contrast to the solid-state greenhouse mechanism proposed by Brown *et al.* (*Science* 251, 1990), melting at the base of a permanent polar cap on Triton caused by internal heatflow can significantly enhance viscous spreading, as well as providing the necessary energy,' fluids, and/or gases to drive Triton's geyser-like plumes; (5) The atmospheric collapse predicted to occur on Triton in the next 20 years (Spencer, *Geophys.*

*Res. Lett., 1991)* **may be plausibly avoided because of the large latitudinal extent expected for permanent polar caps on Triton.**

## Introduction

Following the identification of methane on Triton by Cruikshank and Silvaggio (1979), Trafton (1 984) addressed long-term volatile transport on Triton. Trafton concluded that deposits of methane ice on Triton could exist in both seasonal and permanent deposits, that these deposits would vary in extent and thickness depending upon the precise phase of Triton's complicated seasonal cycle (Harris, 1984), and that any permanent methane polar caps on Triton should exist poleward of the latitude that received the seasonal-mean insolation (near ± *45 O).* Trafton also argued that the atmosphere of Triton should be in vapor-pressure equilibrium with the methane ice on its surface, and that it should have rather large excursions in pressure and density as the amount of surface ice coverage varied through Triton's seasonal cycle.

With the groundbased identification of nitrogen on Triton, and the implication that condensed nitrogen is ubiquitous on Triton's surface (Cruikshank *et* a/., 1984), the problem of volatile transport on Triton became somewhat more stimulating. Results of more sophisticated modeling, however, were published only after the Voyager flyby of Neptune (Smith *et a/.,* 1989) determined enough of Triton's physical properties to strongly influence ideas regarding volatile transport on Triton.

Voyager found that Triton's surface temperature is about 38 K and its atmosphere is composed almost entirely of nitrogen at a surface pressure of about 15 $\mu$bar--very near the equilibrium vapor pressure of solid nitrogen at 38 K (Conrath *et a/., 1989;* Broadfoot *et aL, 1989;* Tyler *et a/., 1989).* An implication of the Voyager measurements is that Triton's surface is mostly isothermal, that isothermality is maintained primarily by latent heat transport (Leighton and Murray, 1972; Trafton, 1984), and that there are extensive deposits of frozen nitrogen over most of Triton's sunlit hemisphere (Smith *et al.,* 1989; Brown *et al.,* 1991; Cruikshank *et al., 1992).*

The high density of Triton, and the relatively large rock fraction thus implied (Smith *et al., 1989),* implies in turn that Triton has a large internal heat source relative to the amount of insolation it absorbs. As we discuss in more detail below, internal heatflow could have large effects on long-term volatile transport on Triton (Brown *et al., 1991).*

The above facts, and the appearance in Voyager images of what seems to be a bright, massive south polar cap that extends nearly to Triton's equator (Smith *et al.,* 1 989), along with darker, redder regions to the north of the equator, have spawned recent reexaminations of the question of volatile transport on Triton. These studies mostly divide into two areas: those that hypothesize that the seasonal frost distribution on Triton is primarily responsible for the apparently large south polar cap and the global albedo distribution (Type 1); and those that hypothesize that permanent ice deposits are instead primarily responsible (Type 2).

The first of the post-Voyager volatile transport models is of Type 2 and was published by Moore and Spencer (1 990). They hypothesized that Triton's south polar cap is primarily a permanent deposit of nitrogen, maintained by a permanent albedo asymmetry between Triton's northern and southern hemispheres. Moore and Spencer argued that differences in the total amount of absorbed insolation for polar caps of different albedo would transport volatiles toward the brighter cap, eventually exhausting one polar cap of its permanent volatile reservoir, while leaving only seasonal deposits that evaporate entirely during summer. Moore and Spencer recognized that a problem with their hypothesis is that it is difficult to maintain a long-term albedo asymmetry between the northern and southern hemispheres on Triton in the face of seasonal deposition of meter-thick frost layers each winter. To maintain the albedo asymmetry, seasonal frost layers in at least the low-albedo hemisphere would have to

**be transparent enough to have no significant effect on the albedo of the substrate.**

**Although recent studies have argued that *semi-transparent* layers of nitrogen are consistent with the anomalous light scattering properties of large areas north of Triton's equator (Lee *et al., 1991* ), adjacent to these areas are other areas which, in contrast, are very efficient at scattering visual photons, and consequently have the highest known albedos on Triton (Smith *et al., 1989;* Hillier *et al.,* 1990). This demonstrates that at least in some places and at some times on Triton, transient, high albedo deposits can form upon an otherwise dark substrate. Furthermore, recent work by Duxbury and Brown (1 993), as well as the seasonal temperature models of Brown (1 993), indicate that any clear deposits of nitrogen on Triton will eventually become fractured by passage through the $\alpha$-$\beta$ phase transition (Scott, 1976), increasing scattering within these layers, and in turn dramatically increasing their albedo. Consequently, the clear deposits required to maintain the hemispherical albedo asymmetry essential for the Moore and Spencer model, if they can be created by seasonal processes, are not likely to survive other seasonal processes. Thus, we conclude that it is improbable that the *hemispherical* albedo asymmetry required by Moore and Spencer (1 990) could be maintained for much longer than a few seasonal cycles on Triton.**

**Partly to alleviate the problems of the Moore and Spencer (1 990) model, Spencer (1990) argued that permanent ice deposits on Triton do not play a large role in seasonal volatile transport and, consequently, in determining Triton's albedo distribution. Spencer initially assumed that long-term volatile transport on Triton is driven only by absorbed insolation, that frozen nitrogen is the high-albedo material seen in the Voyager images, and that the nitrogen ice is underlain by a considerably darker and involatile, water-ice substrate. Under these assumptions, Spencer's model predicts an albedo distribution on Triton substantially different than that seen: that is, areas predicted to be high albedo are instead low albedo, and vice-**

versa.

In an attempt to reconcile his model with the actual albedo distribution on Triton, Spencer (1 990) hypothesized that fresh, thick (> 1 m) nitrogen-ice on Triton might *instead be dark and underlain by a bright substrate.* In this case, as summer progresses in a given hemisphere, the ice sublimates and thins, becoming increasingly brighter and eventually stabilizing itself against further sublimation by virtue of its high reflectivity. Conversely, as the nitrogen ice layer thickens in areas where condensation is occurring, it gets darker. Energy balance calculations by Stansberry *etal. (1* 990) suggest that condensation of nitrogen ice should presently be occurring over most of Triton's northern hemisphere. Under these conditions Spencer's dark-frost model agrees with what is seen: that Triton's northern hemisphere is mostly dark while its southern hemisphere is mostly bright.

Although Spencer's (1 990) dark-frost hypothesis is appealing in its simplicity, it has problems. The first is that it is difficult to find a mechanism that would allow freshly condensed nitrogen frost to be *intrinsically* dark rather than bright. " Anticipating this difficulty, Spencer (1 990) argued that dark particles formed in Triton's atmosphere could be entrained in the frost as it condenses, thus darkening the condensate. Unfortunately this does not solve the problem for several reasons. First, it doesn't seem possible to produce dark particles in Triton's atmosphere quickly enough to significantly darken a seasonal frost deposit (Pollack eta*l*., *1990) --a* theory which is supported by Triton's presently overall high albedo (Smith *et aL, 1989;* Hillier *et a/.,* 7990). Second, Pollack *et a/. (1* 990) have argued that the haze particles in Triton's atmosphere are in fact brighter than much of Triton's surface. Thirdly, if dark particles actually precipitate from Triton's atmosphere, they would likely be composed of organic polymers, and thus would be involatile at 38 K (Thompson and Sagan, 1990), Unless some other mechanism acts to remove or bury entrained dark particles (such as has been suggested, but not demonstrated for dust particles on the martian polar caps), a lag deposit should form as a frost layer evaporates, darkening rather than brightening the thinning layer.

Another complication in predicting the albedo of freshly condensed frost on Triton accrues from a recent study of the thermophysics of grain metamorphism in nitrogen frost by Eluszciewizc (1 991). His work suggests that, if the initial sizes of the frost grains are ≈ 1 $\mu$m, nitrogen will first condense from Triton's atmosphere as a *bright* frost, and will later evolve to a *c/ear,* polycrystalline layer in a time shorter than an average Triton season (164 years). If this is the dominant condensation process for frost on Triton, a freshly condensed frost layer would be bright; later its albedo would approach that of the substrate as the frost metamorphosed to a transparent layer. If the ice grains are instead tens to hundreds of microns in size initially, they will not metamorphose to a clear layer during a typical Triton season and thus will remain bright. In either case, with a bright substrate, the overlying nitrogen ice layers will appear bright,

More recently, Spencer and Moore (1 992) have modified their volatile transport models to include, among other things, finite thermal inertia for the presumed water-ice substrate on Triton. They conclude that subsurface heat conduction can have a significant effect on the seasonal volatile distribution on Triton. In particular, if the thermal inertia of the substrate material (presumably water ice) is significantly higher than about 30'% of that of non-porous water ice at 40 K, Spencer and Moore argue that the boundary of any $N_2$-ice cap in the northern hemisphere would have receded above the terminator at present, and thus would not have been seen in the Voyager images. If the thermal inertia in the northern hemisphere of Triton were about half that of non-porous water ice, then no frost could condense in Triton's northern hemisphere at any time during a seasonal cycle, and Triton would thus be devoid of any north polar cap; in essence, this mechanism would halt hemisphere-to-hemisphere volatile transport on Triton.

Another recent attempt at modeling volatile transport on Triton was made by Hansen and Paige (1 992). They constructed models that are quite detailed and take into account physical effects (primarily the thermal inertia of the frost layer) ignored in previous models. In agreement with Spencer (1990), Hansen and Paige conclude that the "dark frost" hypothesis is most consistent with their models and with the appearance of Triton. In addition, they conclude that if the south polar cap of Triton is indeed composed of $N_2$ ice, it must be a permanent feature, otherwise their models predict that cap's northern boundary would have been much closer to the south pole than was observed.

Yelle (1 992) has also offered a model to explain the lack of bright, seasonal frost in the regions north of Triton's equator. In his model, he advances the hypothesis that frost condensing in Triton's northern hemisphere does so preferentially on north-facing slopes and in areas shadowed by topography. As such, the newly condensed frost contributes little or nothing to the sunlight reflected from these regions and the albedo remains near that of the darker substrate. In contrast to other models, Yelle's model does not have to resort to either dark frost or hemispherical differences in thermal properties to explain why Triton's northern hemisphere is relatively dark. As such, Yelle's model may indeed be preferable to others as regards seasonal volatile transport on Triton because it requires topography at a scale that is clearly evident in the Voyager images. Yelle's model does not deal with the question of the quasi-permanence of Triton's south polar cap, however.

In this paper we pursue different line of inquiry. The primary motivations are: the implausibility of the "dark frost" hypothesis; the ad hoc nature of some previous models; the inadequate investigation of the hypothesis that the south polar cap of Triton is quasi-permanent; and the omission of a major energy source and other important physical effects in previous models. Accordingly, the hypothesis we investigate here is that Triton's south polar cap is primarily a permanent feature, created and maintained by a combination of internal heatflow, the seasonal-mean insolation gradient and the finite viscosity of nitrogen ice. This hypothesis was first advanced by Brown *et al. (1* 991) and explored analytically by Kirk and Brown (1 991a, 1991 b). In this paper we construct a two-dimensional numerical model of Triton's internal heatflow that is coupled to the volatile distribution on Triton's surface by virtue of the large difference in thermal conductivity between nitrogen ice and water ice at the temperatures of interest. We show that, depending upon the total nitrogen inventory and the global distribution of internal heatflow, Triton's south polar cap could be mostly permanent, and that its counterpart in the northern hemisphere could be anywhere from equal in latitudinal extent to non-existent as a permanent feature. In addition, we show that permanent deposits of nitrogen on Triton can be hundreds of meters thick, and that the high temperatures at the base of these deposits induced by internal heatflow may induce basal melting. We speculate that this basal melting might result in geyser-like activity on the south polar cap as warm nitrogen buoyantly rises to the surface of the ice layer.

## Model Formulation

A schematic of the basic model appears in Figure 1. A major assumption is that Triton is completely differentiated. As such it has a rocky core, a water-ice mantle and crust, and a surface layer of volatile ices consisting primarily of nitrogen. For the relative proportions of rock and ice we adopt those derived by Smith *et al. (1* 989) from simple compressional models. The ≈ 70% rock fraction implied by Triton's mean density of 2.06 gm cm-l (Smith *et al.*) yields a core radius of ≈ 1000 km and a mantle and crust of total thickness about 350 km.

To estimate the total inventory of nitrogen on Triton we follow the discussion of Cruikshank *etal. (1984),* but employing the Voyager estimates of Triton's density, if Triton has the solar abundance of nitrogen (roughly about 1 part in 1000), and it is completely outgassed in the form of molecular nitrogen, this would be equivalent to a global layer of solid about 1 km thick. If Triton's surface nitrogen inventory is analogous to that of Titan (about 120 m of solid if Titan's atmosphere were completely condensed, or ≈ 0.08 the solar abundance of nitrogen), the equivalent solid layer would be about 80 m thick. A total inventory of nitrogen on Triton that is much greater than 1 km requires mechanisms that would enhance Triton's nitrogen abundance significantly over that of the Sun, a scenario we do not consider likely. Nor are we inclined to argue that the abundance of nitrogen on Triton could be much greater than that of Titan. Even though Triton likely formed much farther from the Sun where it could have incorporated more nitrogen than did Titan, being smaller and less massive than Titan, Triton may have lost some of its original complement of nitrogen by various processes such as photo-ionization or hydrodynamic escape (Lunine and Nolan, 1992). Thus, we assume that the total inventory of nitrogen ice on Triton is less than 1 km globally, and a nominal value of 100 meters is adopted.

To properly model the long-term volatile transport on Triton's surface one must consider all significant energy sources. Following the work of Brown *et al. (1991)* we consider absorbed insolation and heat from the decay of radionuclides in Triton's core. We neglect the contribution of thermal radiation and scattered sunlight from Neptune because their sum is two orders of magnitude less than absorbed insolation on Triton (Brown *et al., 1991).* The amount of insolation absorbed by Triton is determined by its global average bolometric bond albedo (Hillier *et aL,* 1991) and the instantaneous solar zenith angle integrated over the projected, illuminated surface of Triton. The bolometric bond albedo of nitrogen ice is assumed to be 0.85 and areas free of nitrogen ice on Triton's surface are assumed to have a bolometric bond albedo of 0.6 (Hillier *et al., 199* 1); the emissivity of nitrogen ice is assumed to be 0.6 (Stansberry *et aL, 1990)* while the emissivity of all other materials is assumed to be 0,9. For the subsolar latitude as a function of time we use the approach of Harris (1 984), modified to include the obliquity of Neptune, the inclination of Triton's orbit and the period of precession of Triton's orbital pole as recently redetermined using Voyager data (Jacobson *et al., 1991).* When Triton's rotation is included, the instantaneous solar zenith angle can then be calculated for any point on Triton's surface. Because we are concerned with *permanent* volatile deposits, we use the seasonal-mean insolation and neglect seasonal fluctuations in the sub-solar latitude on Triton. This is acceptable because seasonal deposits are ≈ 1 m thick, while permanent deposits are by assumption tens to hundreds of meters thick.

For the total amount of heat produced in Triton's core by radioactive decay we adopt the value reported by Brown *et al. (1* 991 ) as appropriate to a core of carbonaceous chondritic material. Thus, at the base of Triton's mantle the global-average heatflow H is assumed to be about 6.0 mW m$^{-2}$, which is equivalent to a surface value of 3,3 mW m$^{-2}$.

To calculate the temperature distribution in Triton's water-ice mantle we use the general form of the thermal diffusion equation expressed in spherical coordinates:

$$(1) \qquad \rho(r,\theta,\phi,T)\, C(r,\theta,\phi,T) \frac{\partial T}{\partial t} = \vec{\nabla} \cdot \left[\kappa(r,\theta,\phi,T)\, \vec{\nabla} T\right] + E(r,\theta,\phi,t)$$

Here *p* is density, C is specific heat, T is temperature, *κ is* thermal conductivity, and E is a volumetric power source such as radioactive decay. The space coordinates are r, $\theta$, and $\phi$ while the time coordinate is t.

Because the primary interest here is in the radial and latitudinal dependence of heatflow, and that azimuthal variations in the thickness of the nitrogen layer on Triton are inconsequential on seasonal time scales, the Laplacian in the equation above is expanded in spherical coordinates and then integrated over the azimuthal coordinate ($\phi$), giving:

(2) $$\rho(r,\theta,T)\, C(r,\theta,T)\frac{\partial T}{\partial t} = \frac{1}{r^2}\frac{\partial}{\partial r}\left[r^2\kappa(r,\theta,T)\frac{\partial T}{\partial r}\right] + \frac{\kappa(r,\theta,T)}{r^2\sin(\theta)}\frac{\partial}{\partial\theta}\left[\sin(\theta)\frac{\partial T}{\partial\theta}\right] + E(r,\theta,t)$$

To simplify and speed the numerical calculations, we assume that the thermal parameters and power sources are constant in time and space. We will discuss the limitations of these assumptions later in the context of the discussion of specific choices of parameters. Thus, expanding the derivatives and combining terms we have:

(3) $$\rho\, C\frac{\partial T}{\partial t} = \kappa\left[\frac{\partial^2 T}{\partial r^2} + \frac{1}{r^2}\left(\frac{\partial^2 T}{\partial\theta^2} + \frac{\partial T}{\partial\theta}\cot(\theta)\right) + \frac{2}{r}\frac{\partial T}{\partial r}\right] + E$$

To determine the distribution of temperature in Triton's mantle, we assume that there are no volumetric heat sources in the mantle, and that in steady state all the heat generated in Triton's core is conducted radially upward through the core mantle boundary (that is, the temperature gradient in the $\theta$ direction is $\cong 0$ at the boundary). The problem then simplifies to one of solving for the temperature distribution in a spherical shell having a constant heat flux incident on its lower surface. Thus the lower boundary condition in the radial direction and the upper and lower boundary conditions in the $\theta$ direction are given by:

where $r_0$ is the radius at the core-mantle interface.

The upper boundary condition for the solution of the temperature distribution in Triton's mantle will ultimately be determined by energy balance at Triton's surface and the thermal

$$\frac{\partial T}{\partial r}\Big|_{r=r_0} = \frac{H}{\kappa} \tag{4}$$

$$\frac{\partial T}{\partial \theta}\Big|_{\theta=-\pi/2} = \frac{\partial T}{\partial \theta}\Big|_{\theta=\pi/2} = 0$$

properties of the layer of nitrogen ice that overlies it. This then leads us to a discussion of the equations governing volatile transport, and to a discussion of the assumptions regarding heat transfer between Triton's water-ice crust and the thin layer of nitrogen ice overlying it.

We assume that the surface temperature of nitrogen ice deposits on Triton is maintained at some global average by efficient transport of latent heat in Triton's atmosphere. For this to hold, the flux of latent heat required to maintain global isothermality must not require pressure gradients so large as to induce supersonic winds. This is the Leighton-Murray (1 972) criterion, which presently holds for Triton. As we will later show, if the global nitrogen inventory on Triton is greater than about 2500 g/cm$^2$ ( $\approx$ 25 meters), the latitudinal extent of Triton's permanent polar caps is likely to result in absorption of enough insolation by the caps that Triton will always obey the Leighton-Murray criterion.

The rate of ice sublimation and redeposition for any point on the surface of Triton is governed by energy balance and the global-average temperature of Triton's ice-atmosphere system. The global ice temperature is then given by (Trafton, 1984; Spencer, 1990):

$$\sigma \epsilon T_I^4 \int_a F(\theta)\, da = /[S(0)+ F/((3)]\ F(\theta)\, da \tag{5}$$

Here $\sigma$ is the Stephen-Boltzmann constant, $\epsilon$ is emissivity, F is a binary function that specifies whether a given portion of the surface is covered with nitrogen ice (that is, F can only have values of O or 1), while S is the amount of insolation absorbed at a point on the surface, a is the total surface area of Triton and H is the local heatflow.

The relation governing mass-transfer by sublimation and recondensation can also be derived by assuming energy conservation. That is, the amount of mass gained or lost by a particular unit area of ice is determined by whether there is a surplus or deficit of absorbed energy versus that radiated to space at a given temperature. This energy flux is scaled to mass flux per unit area by the latent heat of sublimation. Dividing by the bulk density $\rho$ gives the rate of change in thickness due to sublimation h, per unit time:

$$\frac{dh_s(\theta)}{dt} = -\frac{s(e) + H(e) - \sigma\epsilon T_l^4}{\rho L} \tag{6}$$

Because the layer of nitrogen ice on the surface of Triton will vary in thickness and coverage as a function of time, we have chosen to simplify the problem of determining the amount of internal heatflow that arrives at the surface of an ice deposit. In particular, because the layer of nitrogen ice on Triton will always be very thin compared to the thickness of Triton's mantle, we neglect any lateral heatflow in the nitrogen layer. In addition, for the purposes of calculating the amount of heatflow conducted from Triton's interior to the surface we assume that on multi-seasonal timescales the average radial temperature distribution in the nitrogen layer is to a good approximation linear, and its slope is determined by the requirement that all heatflow that gets to the base of the layer is conducted to the surface at a rate determined only by the average temperature difference between the top and bottom surfaces of the layer. This assumption greatly simplifies the calculations because it eliminates the need to solve the diffusion equation within a layer whose thickness and latitudinal coverage is time variable. Thus, the upper boundary condition in the radial direction for the solution of the diffusion equation in Triton's mantle is given by:

$$\kappa_{N_2}\frac{\partial\bar{T}}{\partial r}\Big|_{r_1<r\le(r_1+h)} = \kappa_{H_2O}\frac{\partial\bar{T}}{\partial r}\Big|_{r=r_1} \tag{7}$$

Here r, is the radius of the upper surface of the mantle, the bar over the T indicates a seasonal time average, and the thermal conductivities are for water and nitrogen ice.

At the top of the nitrogen-ice layer the temperature is equal to the global-average temperature as defined in and calculated from Equation 5, using the seasonal-mean insolation as a function of latitude. Thus, with this and Eq. 7 above we can immediately calculate the temperature distribution in the nitrogen layer, which is linear with depth. In particular, the temperature at the base of the nitrogen layer and at the top of the water-ice layer it overlies is given by:

(8) $$T(\theta)\big|_{r=r_1} = T_1 + \frac{H(\theta)h}{\kappa_{N_2}}$$

We apply this equation as a boundary condition to that portion of the surface covered with nitrogen, using the global temperature $T_1$ from the previous timestep in the numerical solution, rather than attempting to solve Equations 5 and 8.

Because those areas of the surface free of nitrogen ice do not participate in the global energy balance that causes nitrogen ice to be everywhere isothermal, the boundary condition is derived by assuming local radiative equilibrium. Thus we have:

(9) $$\sigma\epsilon T^4\big|^{\theta}_{r=r_1} = s(\theta) + H(\theta)$$

In the numerical solution, Equation 9 is solved iteratively, using a Newton-Raphson algorithm.

Equations 5-8 describe the feedback loop between Triton's thermal state and volatile distribution, namely the insulating effect of the nitrogen cap; that is, the influence of heatflow on the volatile distribution and the influence, in turn, of the volatile distribution on the heatflow. There is another feedback loop to be considered, however: the viscous spreading of nitrogen ice deposits on Triton.

In previous models it was assumed that any permanent nitrogen deposits would reside very near Triton's poles, with little or no latitudinal extent. If, however, the nitrogen inventory on

Triton is significantly larger than that needed to supply seasonal volatile transport (as we have argued above using Titan as an analogy), solid-state creep, as a mass-transport process that opposes the poleward transport driven by insolation, could be a dominant process in determining the equilibrium thickness and latitudinal extent of any permanent polar caps. With this in mind, the final step in building the model is to specify the scheme by which viscous flow in the nitrogen ice is included. Theoretical models of ice-sheet conformation consist of three parts: (1) a mass-balance equation relating the local rate of change of cap thickness to the net precipitation/sublimation rate and the velocity field that describes the spreading of the cap; (2) a description of the distribution of precipitation and sublimation; and (3) a model by which the velocity may be calculated for a specified cap thickness and slope. We have already described our sublimation/deposition model for Triton, and our velocity model, like those for terrestrial ice sheets, is based on an analogy with glacier flow, which is geometrically simpler and better understood.

The viscous spreading of terrestrial ice sheets has been studied extensively (e.g., Paterson, 1981 ) and forms an obvious analog for the spreading of Triton's polar cap. Terrestrial glaciers flow by two mechanisms. The first mechanism, which always operates to some degree, is a shear deformation in which the upper layers of the glacier ride over the lower ones (Paterson, 1981; pp. 83-89). The second mechanism is basal sliding, which can occur only if the base of the glacier is at its melting point, so that a thin film of lubricating fluid is present (Paterson; pp. 112-1 29). It is not clear whether basal sliding is probable or even possible on Triton. The ambient temperature of 38 K is only 0.6 of the melting temperature of 63.2 K, but, as we will discuss below, the temperature at the base of a kilometer-thick cap on Triton could reach the melting point. A greater uncertainty is whether the thermodynamic properties of nitrogen permit basal sliding, even if a melt layer is present. If the melt layer is thick compared to the underlying topography, the glacier is effectively decoupled from the surface and "surges" at extremely high velocities; this is physically possible on Triton, but terrestrial experience indicates that surges are short lived because the required volume of melt is soon lost .

When the melt layer is thin enough that the glacier contacts the underlying topography, it must flow past the obstacles by a combination of mechanisms. The first is solid-state deformation of the ice around the obstacle, which is most efficient for large obstacles. The second is regulation--pressure-induced melting upstream of the obstacle followed by mass transport in the melt layer and refreezing downstream. This mechanism is most efficient for small obstacles. For some intermediate size obstacles, both mechanisms are equally inefficient and it's these obstacles that control the rate of sliding (Paterson, 1981; pp. 153-184). Nitrogen, unlike water, undergoes pressure-induced *freezing* rather than melting, making regulation impossible. If basal sliding occurs in a nitrogen cap, it must therefore be controlled by solid-state creep around the smallest obstacles present--or at least by the smallest obstacles that penetrate the melt layer and contact the base of the solid.

We have no way of determining either the melt-layer thickness or the small-scale relief beneath Triton's polar caps, and hence no way of quantifying basal sliding. Even so, though we only consider here models for which the basal temperature of the cap is below the melting point, topography seen in the Voyager images of Triton suggest that subsolidus flow may be the dominant mode of flow in the polar caps. In areas only slightly north of the edge of Triton's south polar cap is abundant topography averaging 0.5- 1.0 km in height (Smith *et al., 1989).* In contrast, there is little topography at the ½-1 km scale on the south polar cap itself, except for areas within the cap near its northern boundary, although topography at scales of a few tens to a few hundreds of meters can not be ruled out (cf., Smith *et al.,* *1989;* Soderblom, private communication, 1993). One explanation for this apparent loss of relief on the south polar cap is that the nitrogen ice is several hundred meters thick and has buried most of the underlying topography. Another explanation is that there simply is no major topography on the water-ice substrate of the souther hemisphere, and the nitrogen ice deposits could be either thick or thin. It seems unlikely that the topography similar to that above the northern boundary of Triton's cap stops a short distance south of the cap boundary, and that the entire area underlying the visible portion of Triton's south polar cap is free of topography at the scale of a km. Thus, we assume that at least some of the expected topography in the southern hemisphere of Triton has been buried by a polar ice layer of thickness comparable to the underlying topography.

For the purposes here, a one-dimensional, extensionless glacier model is employed and modified to determine the mass flux due to viscous flow. In this model, viscous flow is driven by the shear stress resulting from the weight of the material above a certain depth in the glacier and the slope of the top surface. Following the approach of Paterson (1 981), but specializing from a power-law rheology to a Newtonian rheology with a temperature-dependent viscosity, we obtain the shear in the flow field as a function of depth z in the layer:

(lo) $$\frac{\partial}{\partial z}u(z,\theta) = \frac{2}{\eta(T)}\rho g \sin(\alpha) z,$$

where $u(z,\Theta)$ is horizontal velocity, g is the acceleration of gravity, $\eta(T)$ is the temperature-dependent viscosity and *a* is the local slope angle of the top of ice layer relative to the horizontal. For small slope angles we can make the approximation that:

(11) $$\sin(\alpha) \approx \frac{1}{r_1}\frac{\partial h}{\partial \theta}$$

Because we have assumed that the flow field depends only on the shear stress from the

**overburden" and the local slope, Equation 10 can be integrated as an ordinary differential equation, for which the appropriate boundary condition is that the velocity must go to zero at the base of the layer. In addition, we make the simplifying assumption that the viscosity is constant and equal to the viscosity at the basal temperature. This is a relatively good approximation because most of the shear occurs near the base of the cap and the upper layers are "rafted" along almost passively. Integrating Eq. 10, then gives the velocity field:**

**(12)**

$$u(z,\theta) = \frac{\rho g (h^2 - z^2)}{r_1 \eta(T)} \frac{\partial h}{\partial \theta}$$

**For a layer of thickness h, we can integrate Eq. 12 to get the velocity of flow in the ice layer averaged over the thickness of the layer:**

***(13)***

$$\bar{u}(\theta) = \frac{1}{h} \int_0^h u(z,\theta)\, dz = \frac{2\rho g}{3 r_1 \eta(T)} h^2 \frac{\partial h}{\partial \theta}$$

**Employing the continuity equation and the fact that the column density is equal to the bulk density times the length of the column, we derive the time rate of change of the thickness of the layer due to viscous flow, dhf/dt:**

**(14)**

$$\frac{dh_f}{dt} = -\vec{\nabla} \cdot \left[ h \, \bar{u}(\theta) \right] = \frac{-2\rho g}{3 r_1^2 \, \eta(T) \cos(\theta)} \frac{\partial}{\partial \theta} \left[ h^3 \cos(\theta) \frac{\partial h}{\partial \theta} \right]$$

**The only thing that remains to complete the formulation is to describe the form of $\eta(T)$. To do so we use an Arrhenius relation:**

**(15)**

$$\eta(T) = \eta_m \exp\left[ A \left( \frac{T_m}{T} - 1 \right) \right] ,$$

**where $\eta_m$ is the melting point viscosity, A is a dimensionless constant, and $T_m$ is the melting temperature. Following Kirk (1990) we adopt the values for nitrogen ice of 8.8 x 1010 Pa-s,**

*24.9* and 63.15 K for $\eta_m$, A and $T_m$, respectively. The values of these parameters, as well as the assumption of a Newtonian rheology, reflect the assumption by Kirk that diffusion creep is the dominant mode of viscous flow for nitrogen ice at the temperatures and stresses of interest. These parameter values are based on solid-state diffusion measurements combined with theoretical arguments. Eluszciewicz (1 991 ) reports rheologic parameters for nitrogen from the literature that were unknown to Kirk (1 990), which lead to similar melting-point viscosities, but which have a stronger temperature dependence. Comparison with his estimates for non-linear creep rates confirms our assumption that diffusion creep is probably the dominant flow mechanism in Triton's polar cap. In any case, the calculated cap thickness and extent for a given insolation depend only weakly on the assumed viscosity law, because the viscosity depends so strongly on basal temperature and, hence, on cap thickness. In essence, Triton's polar cap, like a convecting planetary mantle is likely to be thermally self regulating.

The importance of the inclusion of viscous flow in any model of nitrogen glaciers on Triton is underscored by the idea that Triton's internal heat source is quite large relative to the amount of insolation it absorbs (Brown *et al., 1991).* This is illustrated by a simple calculation. The thermal conductivity of water ice is about 50-100 times that of nitrogen ice at 40 K (Scott, 1972; Klinger, 1980), thus requiring the steady-state thermal gradient in an overlayer of nitrogen to be 50-100 times that in the mantle. Models of the bulk composition of Triton by Smith *et al. (1* 989) indicate that the present thermal gradient in Triton's mantle (assuming a chondritic abundance of radionuclides in Triton's core) is about 0.3 K km-l near the surface. Thus, a km-thick layer of nitrogen on Triton would have a total temperature difference between its upper and lower surfaces of about 15-30 K. Assuming that the top surface of the ice is at the 38 K global temperature of Triton, the temperature at the base of

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First, as previously noted, accumulation of a thick basal melt layer may decouple the cap from its bed, causing it to surge outward, spreading at much higher velocities than are possible for viscous creep. Second, the melt may flow out laterally along the cap base, perhaps breaking out catastrophically in a process similar to that seen in terrestrial glaciers. Melt escaping from under the cap will simultaneously freeze and boil, with the newly produced vapor recondensing rapidly nearby; the overall effect will be a thinning of the cap and transport of nitrogen to the cap edge. Lastly, the liquid nitrogen may escape upwards through crevasses in the cap, because, unlike water ice, liquid nitrogen is much less dense than its coexisting solid, regardless of whether the solid is in the *a* or 13 phase. Again, the net result will be a thinning of the cap and extension to lower latitudes. It is interesting to speculate that vertical migration of liquid nitrogen through fissures in the cap could provide a mechanism for driving Triton's geysers, using internal rather than solar energy. In addition, the locations of these geysers could seem to track the progression of the subsolar point on Triton. This speculation will be explored further later in this paper.

In summary, solid-state creep within a polar ice cap can provide the negative feedback required to prevent the otherwise inexorable, insolation-driven, poleward retreat (Spencer, 1990) of the permanent ice deposits on Triton. This feedback will be intensified by Triton's internal heatflow, which can substantially raise the temperature at the base of the cap, dramatically lowering its viscosity. The numerical models presented below suggest that subsolidus creep is indeed capable of regulating the cap thickness and latitudinal extent. If, however, the viscosity of solid nitrogen is too high (even at the melting point) to offset the insolation-driven retreat of the cap, the increasing thickness of the cap should eventually produce a basal melt layer, the redistribution of which will provide additional, more-efficient, feedback mechanisms for stabilizing the cap. These mechanisms may come into play either if the nitrogen is stiffer than we have assumed or if Triton's volatile inventory is somewhat larger than we have assumed.

Besides the facilitation of viscous flow in Triton's nitrogen glaciers, internal heatflow can also be very important in the long-term transport of nitrogen ice by surface sublimation and recondensation. To illustrate this, consider that the global-average heatflow on Triton is about 3.3 mW $m^{-2}$. If local variations about the global average were only 2.5%, this would represent differences in heatflow at points on Triton's surface of about 0.1 mW m“$^{2}$. Using the latent heat of sublimation L = 2.6x 102 j/g for nitrogen ice (Brown and Ziegler, 1979), the implied differential sublimation rate would be about 3.8 x 1011 g $cm^{-2}$ S-l, or about 12 m per million years! Thus, on geologic time scales, anisotropies in Triton's internal heatflow could have strong effects on the distribution of Triton's *permanent* ice deposits. As mentioned before and demonstrated below, there is a coupling between the spatial distribution of internal heatflow and the nitrogen-ice surface layer, which, in some cases, can maintain and magnify the original anisotropies.

## Numerical Solutions and Specific Cases

In this section we discuss the results of calculations of specific cases applicable to Triton using the formulation derived in the previous section. Before we proceed with that, however, it is useful to briefly describe the numerical techniques used to solve the coupled set of equations derived above.

Solutions of the diffusion equation (Eq. 3) were accomplished using a fully implicit finite-differences formulation (e.g., see Press et *al.*, 1988), generalized to multiple dimensions. To minimize the number of numerical operations required to solve the diffusion equation for a multi-dimensional spatial grid, the Alternating Direction Integration technique was applied. This technique allows one to decouple integrations in different spatial directions by breaking up the numerical time step into equal parts, inverting the finite difference matrices separately for each spatial direction and making use of the temperature field derived from the previous fractional time step. The primary advantage of this technique is that the finite-difference matrices are always tridiagonal, thus dramatically speeding the process of inverting the matrices and solving for the temperature field, with little or no loss in accuracy.

The upper and lower boundary conditions in the radial and $\theta$ coordinates were specified as described by Eqs. 4 and 8 above. The boundary conditions in the $\theta$ coordinate results from the assumed azimuthal symmetry of *our* model.

Because we are concerned here with volatile transport on time scales much longer than a typical seasonal cycle on Triton, we use the seasonal mean insolation (Brown *et al., 1990)* and a time step that is relatively large, typically 250-1000 years. The spatial grid used was 40 by 60 in the r and $\theta$ directions, respectively, *representing* a reasonable tradeoff between accuracy and execution time when solving the diffusion equation.

Integration of the mass-transport equations (Eqs. 6 and 14) was accomplished with a simple Euler integration, executed at each full time step in the solution of the diffusion equation.

Because both mass transport relations must be mass conservative, at the end of each time step the total nitrogen inventory was calculated and compared with the initial inventory to ensure that round-off errors did not accumulate.

The general procedure used in the computer code consisted of advancing the solution through each time step by first solving for the temperature field in the mantle using the initial conditions from the previous time step and the 4 boundary conditions. After the mantle temperature field was determined, surface heatflows were calculated, the global ice temperature was determined and the mass flux due to sublimation and viscous flow calculated. The total amount of ice over a given point on the surface was then incremented or decremented appropriately, mass conservation checked and the whole process repeated at the next time step. In the cases we detail below, we will describe the specific parameters used, but a single run of the model usually lasted for 50 million machine years to ensure that the final solution was independent of initial conditions.

As was mentioned previously, we have assumed that the thermal diffusivity of Triton's mantle material is independent of temperature and location. The primary reason for this simplifying assumption is to reduce the execution time of the computer code. It is clear that the thermal diffusivity of water ice is indeed dependent upon temperature, but for the cases we consider here, we are not so much interested in the exact temperature distribution of Triton's mantle, as we are in the approximate magnitude and direction of the thermal gradients and thus the internal heatflow. As such, we have assumed that the thermal conductivity of our water-ice mantle is $10^6$ erg see-l $cm^{-1} K^{-1}$, the heat capacity is 107 erg gin-l (Giauque and Stout, 1936) and the density is 1 gm $cm^{-3}$ (Scott, 1976). Because we are interested here in the steady-state thermal gradients inside Triton, it is only the thermal conductivity that is of importance as regards the thermal parameters. As such, the value cited above represents the thermal conductivity of water ice at about 80 K (Klinger, 1980), roughly the expected radially averaged temperature for the heatflows associated with a chondritic core. Because the thermal conductivity of water ice falls with rising temperature, the effect in our models of using a constant conductivity is to underestimate slightly some of the effects due to lateral heat transport within Triton's mantle.

It should be noted at this point that our model allows neither for phase changes nor for . convection within Triton's water-ice mantle. Neither process is likely to occur if the global heatflow is 3.3 mW $m^{-2}$, as we have assumed. With this heatflow, and our assumed conductivity, the temperature at the base of the mantle is about 20 K below the $H_2O$-$NH_3$ peritectic and about 120 K below the melting point of pure water. Kirk (1991a) performed a convective-stability analysis that shows that the water-ice layer on Triton is too thin to convect with this temperature gradient and any plausible viscosity law. if, however, the heatflow on Triton is much higher than we have assumed, the thermal gradient will be much steeper, and some deep seated melting may be possible accompanied by subsolidus convection above the melting layer, if the viscosity of the ice is relatively low (Kirk, 1991a). Either of these phenomena will increase the efficiency of lateral heat transport, but both will be limited to the lower portion of Triton's water-ice mantle. Heat transport in the overlying ice will still be conductive, and our model will still be valid, provided that the lower boundary condition is applied at a shallower depth. The long-term consequences of the feedback between the conductive thermal model and 'the volatile distribution will be qualitatively unchanged, but we expect that the redistribution of volatiles will take place more quickly than if the entire mantle were conductive.

*Case 1: Uniform Insolation*

As was mentioned in the previous section, one of the most important results of this work is the recognition that internal heatflow on Triton and its surface volatile distribution can, in some cases, strongly couple. To demonstrate how this coupling occurs, we first perform a simple thought experiment.

In Figure 2 is an illustration of a section of the surface and interior of a model body that has a uniform internal heat source, a water-ice mantle, and a surface layer of volatile ice, in this case nitrogen. Assume for the moment that the initial distributions of temperature and heatflow in the mantle of our hypothetical body are laterally isotropic with a temperature that increases with depth, as are the temperatures at the upper and lower surfaces of the nitrogen ice layer. it is also assumed that above the nitrogen ice is an atmosphere that is capable of transferring latent heat such that the surface of the layer is isothermal. As can easily be seen, the initial condition of lateral isotropy in the temperature distribution of the mantle cannot hold at any later time because the nitrogen layer has random topography. To maintain isotropic heatflow through the nitrogen-ice layer to its surface, the temperature at the base of the layer must be such that it is higher below areas that are topographical highs and lower below areas that are topographical lows. This, of course, will force lateral heatflow in the surface and interior of the mantle from areas of high temperature to areas of low temperature, ultimately resulting in an anisotropic heatflow field. For the moment we will not worry about lateral heat transport within the nitrogen ice layer itself because it is small compared to that in the mantle by virtue of the large difference in thermal conductivity between water and nitrogen ice.

The net effect will be that heatflow within the mantle will be such that a small amount of heat will be driven laterally from areas that are topographic highs to areas that are topographic lows. If we allow that the atmosphere above the ice can efficiently transport latent heat, net mass transport from the thinner areas to the thicker areas will occur, and the system becomes unstable. Initially the layer will segregate into patches of ice, made thicker by accretion of transported material. These patches of ice will then compete until only one containing the entire mass of the original ice layer remains; i.e, survival of the thickest. If the surviving patch has infinite viscosity, differential sublimation driven by the differing heatflows across its extent will eventually cause the patch to become infinitely high with *no* areal extent. For real materials, however, this will not occur because the layer will sag of its own weight and mass will flow laterally, eventually balancing the differential sublimation and stabilizing the layer at a finite height and areal extent.

This spontaneous instability in a non-uniform, volatile ice layer in the presence of an internal heat source is demonstrated in Figure 3, where is illustrated a run of the model for a hypothetical body that has the configuration of Fig. 1. To demonstrate the coupling between internal heatflow and the volatile distribution, the body is uniformly illuminated with sunlight at the 1'.5 W $m^{-2}$ level available at 30 AU from the sun, The body also has the size and density of Triton, and an initial volatile inventory of 50 meters of nitrogen, uniformly distributed except for random topography of ± 10% of the average thickness of the layer. For initial conditions, it was assumed that the temperature gradient in the mantle was purely radial and in equilibrium with an internal heatflow amounting to 6.0 mW $m^{-2}$ at the core-mantle interface. The bolometric bond albedo was assumed to be 0.85 in areas covered with nitrogen ice. Areas bared of nitrogen ice expose water-ice with a bolometric bond albedo of 0.6 (Hillier *et aL,* 1991).

In Fig. 3 is plotted the thickness of the ice layer as a function of time and latitude, for a period of 20 million years. At t =0, the layer is 50 ± 10 m in thickness globally and the temperature field in the mantle is laterally isotropic with a radial gradient such that the system is in steady state with a surface heatflow of 3.3 mW $m^{-2}$. As can be seen from the figure, within the first 2 million years or so, the ice layer experiences its first "burn throughs". That is, the areas that are initially thinnest, evaporate entirely down to the water-ice substrate. The remaining areas compete over the next several million years with the thickest areas always growing thicker at the expense of the thinnest areas. By the time 10 million years have passed most patches have evaporated away and have been incorporated into the two thickest areas which by then have become several hundred meters thick. The final two areas compete until there is only one left 16 million years from the start of the simulation, and the surviving annulus has only viscous spreading to modify its thickness profile after it has gobbled up all its competitors. That can be seen in Fig. 3 as a slight increase in latitudinal extent with an accompanying reduction in maximum thickness. It is also useful to note here that the final nitrogen deposit lies in a band between 30° and 60° south latitude. As such, it is quite extensive and demonstrates that viscous spreading can successfully counteract the tendency of the deposit to become thicker and less extensive with time.

There is nothing significant, however, about the central latitude of the final deposit in our illustration because it depends on initial conditions; that is, if we started with a slightly different volatile distribution, we would end up with the final deposit at some other central latitude. In addition, since we have enforced azimuthal symmetry, the final result in a fully three-dimensional case would not bean annulus, but rather a cap, because the ice layer would be unstable in longitude as well as latitude. This spontaneous breakup of a volatile layer due to the action of internal heatflow may be significant for bodies for whom the long-term average insolation is minimized at the equator instead of at the poles.

As was mentioned before, the gradient in heatflow across the surviving patch is driven by the thickness profile of the patch. This is shown in Figure 4, in which the average heatflow reaching the surface of mantle below the ice layer is displayed as a function of latitude. As can be seen, the heatflow in the interior of the region is less than that at the edges, delivering more energy to the edges and transporting mass from the edges to the center. This mass flux is opposite in direction to that to that of viscous spreading, eventually stabilizing the patch in latitudinal extent.

The spontaneous breakup of a volatile layer in the presence of internal heatflow may be applicable to the case of Pluto. The long-term minimum in the average insolation on Pluto lies at its equator instead of at both poles (Dobrovolskis and Harris, 1983) as it is for Triton. As we will see in the case cited below, permanent volatile deposits like to reside at long-term insolation minima, and thus should exist at Pluto's equator, rather than its poles, resulting in a permanent equatorial ice "band". If there were any azimuthal anisotropies in either the thermal conductivity of this ice band or in its thickness, the resulting asymmetric heatflow would cause a spontaneous breakup of the equatorial ice "band" eventually producing an ice cap on Pluto's equator. The expected azimuthal orientation of this cap would be random, except in the case that there is another energy source in the system that can force a preferred location. Because the Pluto-Charon system is thought to be tidally locked, scattered sunlight and radiated thermal energy from Charon may provide a preferred location for an equatorial ice cap on Pluto. In a follow-up paper, we will generalize the model used here to three dimensions and explore this speculation in detail.

*Cases 2 and 3: Realistic Insolation With and Without Forced Asymmetric Heatflows*

To illustrate the volatile migration effects under conditions that are more apropos to a body like Triton, we have calculated two additional models, both having the value of the long-term mean absorbed insolation calculated using the best-available values of Triton and Neptune's orbital parameters (Jacobson, 1990).

The first of the two model varieties is shown in Figure 5, where is plotted the depth of the nitrogen layer as a function of latitude and time, assuming an initial depth of 100 meters and initial topography of 10%. In this model we have used the mean insolation as a function of latitude for Triton as calculated by Brown *et al. (1* 990), and assumed that the heatflow at the base of Triton's mantle is uniform and equal to a surface value of 3.3 mW $m^{-2}$; the albedos of frost and substrate are as previously specified. As can be seen from the *figure,* the initially global nitrogen layer is quickly driven away from the equator, and within about 1 million years, the entire volatile inventory is sequestered in two polar caps, each having a thickness of about 1 km at the poles and extending to about ± 550, respectively. The edges of the polar caps continue to retreat for several million years before the center of the cap becomes thick enough to drive significant viscous flow toward lower latitudes. Eventually the edges of the caps stabilize at latitudes of about ± 450. The lowest latitudes at which the polar caps will be stable depends on the total inventory of nitrogen, the global-average heatflow, the temperature dependent viscosity of solid nitrogen and the size of the insolation gradient. Greater heatflow or a larger total nitrogen inventory will result in greater areal extent, as will lower viscosity and a shallower insolation gradient. With nominal values for heatflow, nitrogen inventory and viscosity, it is clear that extensive *permanent* polar caps are quite plausible for Triton, suggesting that permanent polar caps may play a large role in controlling both the albedo distribution and volatile transport processes on Triton. As such, consideration of only seasonal volatile transport on Triton is likely to be an oversimplification of the problem. Although the volatile distribution forces variations in Triton's global heatflow and visa versa, other forced heatflow asymmetries may occur on Triton. For example, if there are convection cells in Triton's core, asymmetries in global heatflow may result. If these heatflow asymmetries are large and extensive enough, they may in fact modify the global distribution of nitrogen ica that would result otherwise from the long-term, equator-to-pole insolation gradient. In other words, the tendency of the insolation gradient on Triton to form permanent caps of nitrogen ice on both poles could be strongly affected, depending on the size and placement of heatflow asymmetries. To illustrate this, we show a final case that is identical to the previous one except that we have introduced a hotspot at the core-mantle boundary. Here the heatflow enhancement amounts to twice the global average value and extends from 450 north latitude to the north pole. As can be clearly seen and expected, Figure 6 demonstrates that the extra energy from internal heatflow available to the north cap (illustrated in Figure *7)* results in net mass transfer to the south cap. The cap-to-cap mass transfer is counteracted by the tendency of the more massive cap to spread to lower latitudes, thereby increasing the total energy available to the cap, eventually halting any further net cap-to-cap transport.

Thus, it is obvious that a number of solutions can be obtained, ranging from two symmetric caps to only one cap, by simply adjusting the amount of internal heatflow available to each cap. Viscosity also plays an important role, and with our present state of ignorance regarding the intrinsic properties of solid nitrogen, as well as the properties of Triton's polar caps, the actual viscosity can be uncertain by as much as two orders of magnitude. Finally, as we have mentioned before, the total inventory of nitrogen on Triton will play a pivotal role as well in determining the ultimate thickness, latitudinal extent and latitudinal asymmetry of Triton's polar caps.

To estimate some rough limits to the latitudinal extent of a theoretical south polar cap on

Triton, the model was run for cases encompassing plausible estimates for the upper and lower limits of the nitrogen melting-point viscosity, the total nitrogen inventory, and the size of the north-south heatflow asymmetry. The respose of the model to a variation of ± 2 orders of magnitude in the melting-point viscosity of nitrogen resulted in about a ± 150 difference in the latitude of the cap edge relative to our nominal case (case 2); that is, the cap edge moved between 60° and 300 south latitude. The major effect of varying the heatflow asymmetry from O to a factor of 2 of the northern hemisphere relative to the southern hemisphere results in a steady retreat of the edge of the north cap as the asymmetry increases, until at about a factor of 1.9 the north cap disappears entirely. Variation of the total volatile inventory from the nominal 100 meters to 1 km resulted in both cap edges varying from their nominal 450 to all the way to the equator at a volatile inventory of 1 km. Obviously, the Voyager images do not show a north polar cap that extends below about 450 north latitude, so without a forced heatflow asymmetry to reduce the latitudinal extent of the north cap, volatile inventories in our model of greater than about 100"meters globally are inconsistent with the Voyager images. If we postulate that a factor of 2 heatflow asymmetry is the largest that is plausible on Triton, then volatile inventories as high as about 200-300 meters could still be consistent with the edge of the north polar cap being above 450 on Triton.

## Discussion

Although, the hotspot model described above does not in and of itself argue conclusively that the distribution of internal heatflow on could be having a large effect on its *permanent* volatile distribution, the surficial evidence for massive and widespread past geologic activity in Triton's northern hemisphere (Smith *etal., 1989;* Strom *etal., 1990;* Shenck, 1993) argues that this is not only a viable model, but perhaps the preferred model.

Part of the rationale for the choice of the case illustrated in Fig. 6 is to demonstrate an end member of models that have large-scale, forced heatflow asymmetries. Another part of the reason for choosing the parameters in this way is to illustrate a mechanism capable of producing asymmetric, permanent polar caps that is different from the mechanisms advocated by both Moore and Spencer (1 990) and Spencer and Moore (1 992).

As the reader will recall, Moore and Spencer (1 990) argued that the albedo distribution on Triton is such that the northern hemisphere is permanently darker than the southern hemisphere, thus creating a cold trap for volatiles in the southern hemisphere. The ultimate result would be a large, permanent south polar cap containing the bulk of the nitrogen on Triton. As previously mentioned, three problems with this hypothesis are: that it is not likely that the required long-term, north-south albedo asymmetry on Triton could be maintained; and that the asymmetry either predated the outgassing of the bulk of Triton's volatile inventory or that some event must have initiated an asymmetry in the volatile distribution which was then maintained by an albedo asymmetry in the substrate.

The hypothesis of Spencer and Moore (1 992), also suffers from the problem of requiring an event to initiate an asymmetry in the volatile distribution which is then maintained, as well as requiring that the water-ice substrate in the near surface of Triton have a thermal inertia substantially higher than that expected for ordinary water-ice regoliths. An alternative, they speculate, is that to initiate and maintain an asymmetry, a large-scale difference in the thermal inertias of the northern and southern hemispheres would suffice, if the initial nitrogen inventory were low enough that the seasonal thermal wave could penetrate the nitrogen layer to the water-ice substrate. As such, the hypothesis of Spencer and Moore is not entirely implausible, but it is somewhat ad hoc and presently untestable because there are no existing or planned measurements for Triton that bear on the question.

In contrast, there is clear indication in the Voyager images of geologically recent, large-scale, crustal volcanic activity over most of the northern hemisphere of Triton. In particular, the are large basins that are likely volcanic flood plains that are hundreds of kilometers in size (Helfenstein *et al.*, 1992), as well as the "cantelope" terrain that covers a large part of the visible northern hemisphere. The unique appearance of the "cantelope" terrain has been attributed to explosive vulcanism (Smith *et al, 1989;* Strom eta*l*., 1990) or extensive crustal diapiric activity driven by high internal heatflow (Shenck, 1992). In addition, there is almost no evidence of equivalent crustal volcanic activity in Triton's southern hemisphere (assuming that Triton's geyser-like plumes are not a manifestation of crustal volcanic activity, but rather a manifestation of processes within its nitrogen ice cap). This suggests the distinct possibility that heatflows in Triton's northern hemisphere have recently been substantially higher than the global average, favoring the current hypothesis.

The hypothesis outlined here is also favored because any large-scale, initial heatflow asymmetry, either conductive or convective, is followed by a rapid rearrangement of volatiles. As we have argued earlier, heatflow asymmetries as small as 0.1 $mW\,m^{-2}$ can move several meters of nitrogen in $10^6$ years. Thus, the initiating event need only have occurred in the last few tens to hundreds of millions of years. The massive geologic activity in Triton's northern hemisphere seems to have occurred within the last few hundred million years (Smith *et al., 1989),* and thus it is quite plausible that the distribution of permanent volatile deposits on Triton has changed substantially throughout Triton's history, depending on the location and severity of past geologic activity, and may even have done so within the last few tens to hundreds of millions of years.

Spencer and Moore (1 992) argue that the presence of isolated, high albedo patches near the edge of Triton's south polar cap "is difficult to explain within the context of the model presented here because all nitrogen deposits must be connected by physically flowing glaciers. They also argue that the presence of a global, linear ridge network in Triton's extreme southern latitudes indicates that any nitrogen ice deposits there are less than 1 km thick. As we argued previously, the edge of the permanent cap in the southern hemisphere is not likely to extend below about 30° south latitude, the bright, isolated patches referred to by Spencer and Moore may in fact be involatile material that is not connected to the south polar cap. The suggestion that these deposits may be composed of $CO_2$ ice is attractive and supported by recent measurements of $CO_2$ absorption in Triton's global reflectance spectrum (Cruikshank *et al., 1993).* Alternatively, although it has not been demonstrated here, the possibility remains that Triton's south polar cap occasionally "surges" due to accumulation of melt at is base, leaving portions of the south polar cap detached from the main cap. If these "islands" are thick enough, they could easily survive for a full seasonal cycle or longer. It is also possible that both types of islands exist: involatile $CO_2$ and volatile nitrogen, Finally, the putative visibility of a global ridge system on Triton in the extreme southern latitudes presents no difficulty for the viscous-spreading hypothesis because our model predicts that the south polar cap only approaches 1 km in thickness at the pole, while rapidly thinning northward.

Another, potentially very interesting, consequence of the high temperature at the base of a permanent cap induced by internal heatflow is the fact that nitrogen ice and/or liquid at the base of the cap, being hotter, will be less dense and thus buoyant, thus possibly leading to buoyant nitrogen rising through fissures within the cap and breaking out at the surface. Recent work by Duxbury and Brown (1 993), suggests that there is a strong phase stratification with depth in Triton's polar caps, with as many as 2 $\alpha$-ß nitrogen phase fronts at any one time propagating downward into the caps to depths of 20-30 meters. This phase stratification may enhance cracking in the upper layers of the cap due to the large density difference between the *a* and ß phases of solid nitrogen. If this form of volatile migration exists on Triton, it could manifest itself in many ways, but "one possible way is that warm parcels of nitrogen ice or liquid will rise to the surface of the cap where the enhanced temperatures could give rise to the gas pressures required to drive Triton's geyser-like plumes.

Because the active plumes on Triton have a finite lifetime and their locations seem to be correlated with the present sub-solar point, and because Triton's streaks seem to have followed the path of the sub-solar point through the latest season (Smith *eta/., 1989;* Hansen *et a/.,* 1990), previous investigators have postulated a solid-state greenhouse mechanism as the driving energy source (Brown *et a/., 1990;* Kirk *et a/., 1990).* This hypothesis has received relatively wide acceptance because it explains the latitudinal distribution and relatively short lifetime of Triton's plumes, as well as some aspects of the active plumes observed by Voyager during the flyby of the Neptune-Triton system (Soderblon *eta/., 1990).* These aspects notwithstanding, it is the seeming correlation of Triton's geyser-like plumes with insolation on Triton that provides that greatest impetus for acceptance of theories that invoke insolation as providing the driving energy for Triton's plumes. This correlation may be misleading, however, because volatile transport models all predict that the *distribution* of volatiles on Triton will be at least partially correlated with insolation. Thus, it is possible that another mechanism, such as the one we have cited above, could be providing the energy necessary to drive Triton's plumes, and that the seeming correlation of Triton's plumes with solar input is simply the result of that fact that the plumes occur where the volatiles are, and the volatile deposits are correlated with insolation.

## Summary and Conclusions

We have constructed a two-dimensional model of conductive heatflow in Triton's interior combined with volatile transport and viscous flow in the nitrogen ice deposits on Triton's surface. We have shown that the distribution of nitrogen on Triton's surface and the distribution of heatflow in Triton's interior can be strongly coupled. This coupling is a result of the large difference in thermal conductivity between the nitrogen ice on Triton's surface and its water-ice mantle. In essence, thick nitrogen deposits on Triton, being poor conductors of heat, have elevated temperatures at their bases which drives heatflow laterally away from the centers of the deposits toward their edges. If the global temperature of Triton is such that latent heat can be efficiently transported from areas with greater heatflow to areas with less, net mass transport will occur, thickening the deposits in areas of lower heatflow, thus reinforcing the effect.

In ice deposits of infinite viscosity, this effect will cause the deposits to be increasingly thicker at their centers while reducing their lateral extent, eventually becoming infinitely thick. When finite viscosity is taken into account, the tendency of nitrogen-ice deposits to sag of their own weight, amplified by the strong, radial thermal gradient induced by heatflow, causes the deposits to eventually sag at a rate that just counteracts the net sublimative transport effected by the horizontal heatflow gradient.

Viscous flow will also tend to counteract the effect of the gradient in mean absorbed insolation on Triton, in all cases resulting in polar caps of finite latitudinal extent. If the initial inventory of nitrogen on Triton is in the range of 100 meters globally, the bulk of the nitrogen on Triton's surface is in permanent polar deposits, extending nearly halfway to Triton's equator.

If a forced heatflow asymmetry is introduced into the system, say from a core convection cell on Triton, depending upon the size and location of the asymmetry, one permanent polar cap can be completely sublimated away and incorporated in the surviving cap. The resulting rapid redistribution of nitrogen ice, by virtue of its thermal-blanketing effect, will reflect and reinforce the new distribution of heatflow in Triton's mantle, in turn reflecting and maintaining the permanent ice distribution. As such, this mechanism is preferable to the mechanism proposed by Moore and Spencer (1 990) to account for the apparent large size of Triton's south polar cap and the apparent lack of a north polar cap.

A further consequence of internal heatflow conducted to the surface through thick deposits of nitrogen in Triton's polar regions is the possibility of basal melting in the cap, with the implication that buoyant parcels of warm liquid or solid" nitrogen could rise to the surface of the cap through fissures to manifest themselves as geysers. A detailed investigation of this possibility will be the subject of a future paper.

Finally, if Triton's south polar cap is permanent and hundreds of meters thick, it will never completely sublimate away during a seasonal cycle, thus possibly preventing the atmospheric collapse predicted by Spencer (1 990) on the basis of models that considered only the seasonal distribution of nitrogen ice on Triton. A test of this hypothesis is possible in the next couple of decades because Triton is currently approaching summer solstice in the southern hemisphere.

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## Figure Captions

**Figure 1: Schematic diagram of model geometry. Triton is assumed to be totally differentiated into a rocky core, a water-ice mantle and crust, and a thin layer of $N_2$ ice, no more than 1 km thick.**

**Figure** *2:* **Schematic diagram illustrating an instability driven by heatflow through a non-uniform layer of $N_2$ ice overlying a layer of water ice. The initial conditions, at t = O, have uniform heatflow transported radially upward through both layers, the upper surface of the nitrogen ice being held isothermal at T, by transfer of latent heat, and the temperature gradient in the nitrogen linear, purely radial and everywhere equal; the slope of the gradient is determined by the ratio of the thermal conductivities of water and nitrogen ice. At some time later, t = At, the temperature distribution will adjust such that the temperature at the base of areas where the layer is thinnest ($Ar_1$) will be lower than at the base of areas where the nitrogen is thicker ($\Delta r_2$). This will drive a small amount of heatflow laterally from thicker areas to thinner areas, magnifying the effect, and transferring mass from the thinnest areas to the thickest. The process will be completed when only one patch containing all the nitrogen remains.**

**Figure 3: Volatile transport model with uniform insolation. Plotted here is the thickness of a global nitrogen layer as a function of time and latitude for a model body with the physical characteristics of Triton. The initial conditions are a global layer 50-m thick on average with ± 5 m of random topography, global heatflow equal to that resulting from a chondritic core**

**of 1000 km radius, and insolation that is uniform with latitude and longitude such that the initial global temperature is 37 K. The model was run for 50 million machine years. As can be seen, this figure demonstrates the instability driven by internal heatflow in a non-uniform volatile layer. As time progresses in this model, thinner areas lose material to the thicker areas, eventually sublimating away completely, with the ultimate result being an annulus of nitrogen concentrated between roughly 20° and 40° north latitude.**

**Figure 4: Plot of basal temperature versus latitude for the remaining annulus in Figure 3. Note that the lateral temperature gradient causes the basal heatflow to be lowest at the center of the annulus and highest at its edges. This drives mass transfer from the edges of the annulus to its center, causing it to grow thicker at the expense of its latitudinal extent. The process is opposed by the tendency of the annulus to flow outward via solid-state creep, eventually stabilizing when the rate of creep equals the recession rate of the edges driven by differential sublimation.**

**Figure 5: Volatile transport model with realistic insolation and uniform heatflow at the core-mantle interface. Plotted here is the thickness of a global nitrogen layer as a function of time and latitude for a model body with the physical characteristics of Triton. The initial conditions are a global layer 100-m thick on average with ± 10 m of random topography, global heatflow equal to that resulting from a chondritic core of 1000-km radius, and the mean insolation as a function of latitude on Triton (Brown *et al., 1991),* such that the initial global temperature is 37 K. As can be seen, this figure demonstrates that viscous spreading can indeed halt the recession of Triton's permanent polar caps such that they can extend down to low latitudes, in this particular case to latitudes of ± 48°. The actual latitudinal extent**

depends, among other things, on the initial nitrogen inventory. The greater the initial volatile inventory, the greater the latitudinal extent of Triton's permanent polar caps.

Figure 6: Volatile transport model with realistic insolation and a hotspot at the core-mantle interface. Plotted here is the thickness of a global nitrogen layer as a function of time and latitude using the same parameters as the plot in Figure 5, except that a hotspot has been introduced at the core-mantle interface such that the resulting heatflow is about 2.5 times the global average expected from a chondritic core (see Figure 7 for a plot of the resulting surface heatflow distribution). As can be clearly seen, an asymmetry in heatflow at this level can result in strong differential sublimation between the north and south caps such that the north cap disappears as a *permanent* feature. In this case, as in the one presented in Figure 5, the remaining cap extends down to about -400 latitude.

Figure 7: Surface heatflow for the model calculation in Figure 6. Plotted here is a snapshot at $t = 2.5$ Myr of the surface heatflow as a function of latitude for the model calculation shown in Figure 6. The hotspot extends from 450 to 90° north latitude at the core-mantle boundary and in magnitude is about 2.5 times the global-average value expected for a chondritic core. The discontinuities at about -50 and +550 latitude indicate the edges of the both polar caps, and result partly from the fact that the latitudinal grid was divided into 2° bins.

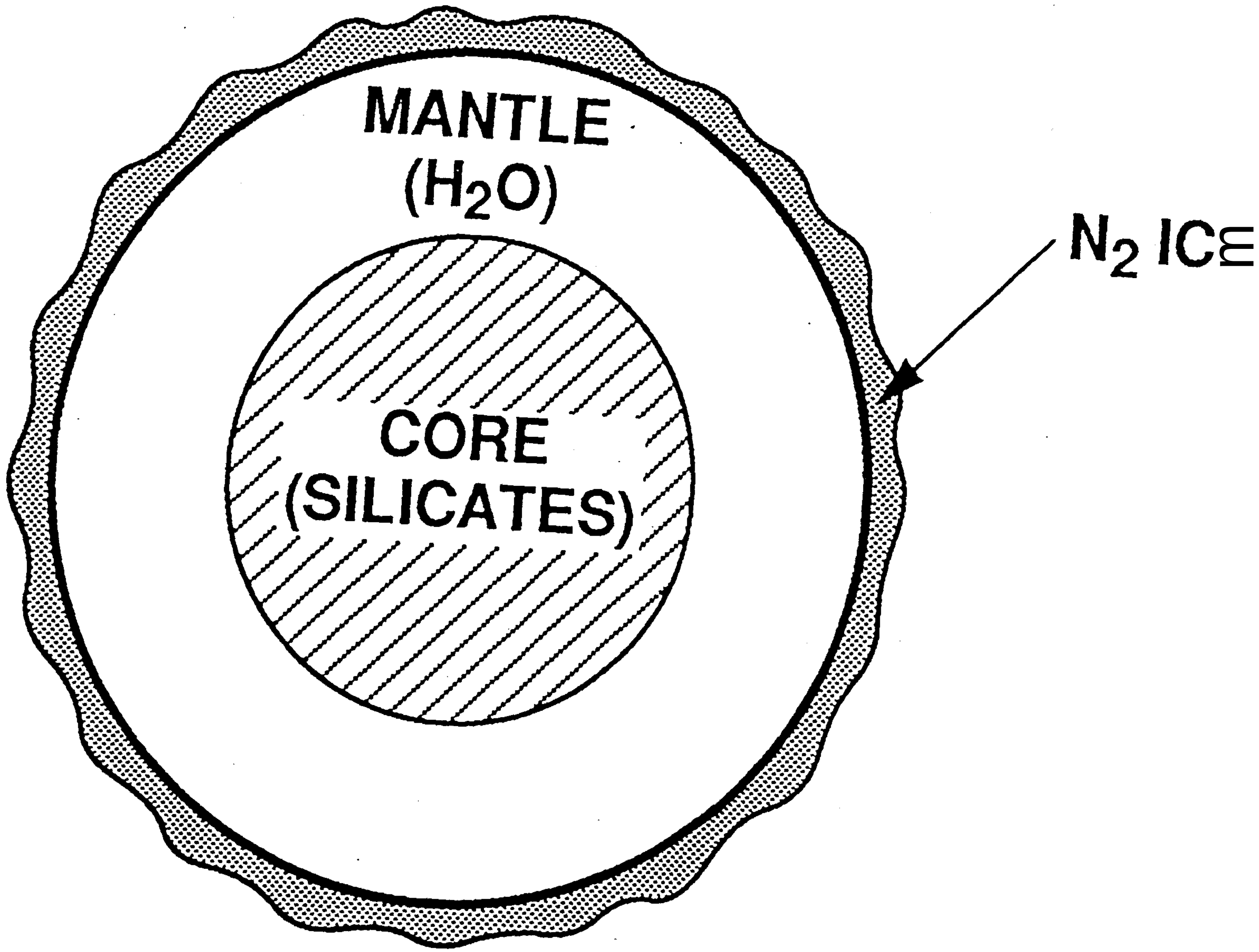
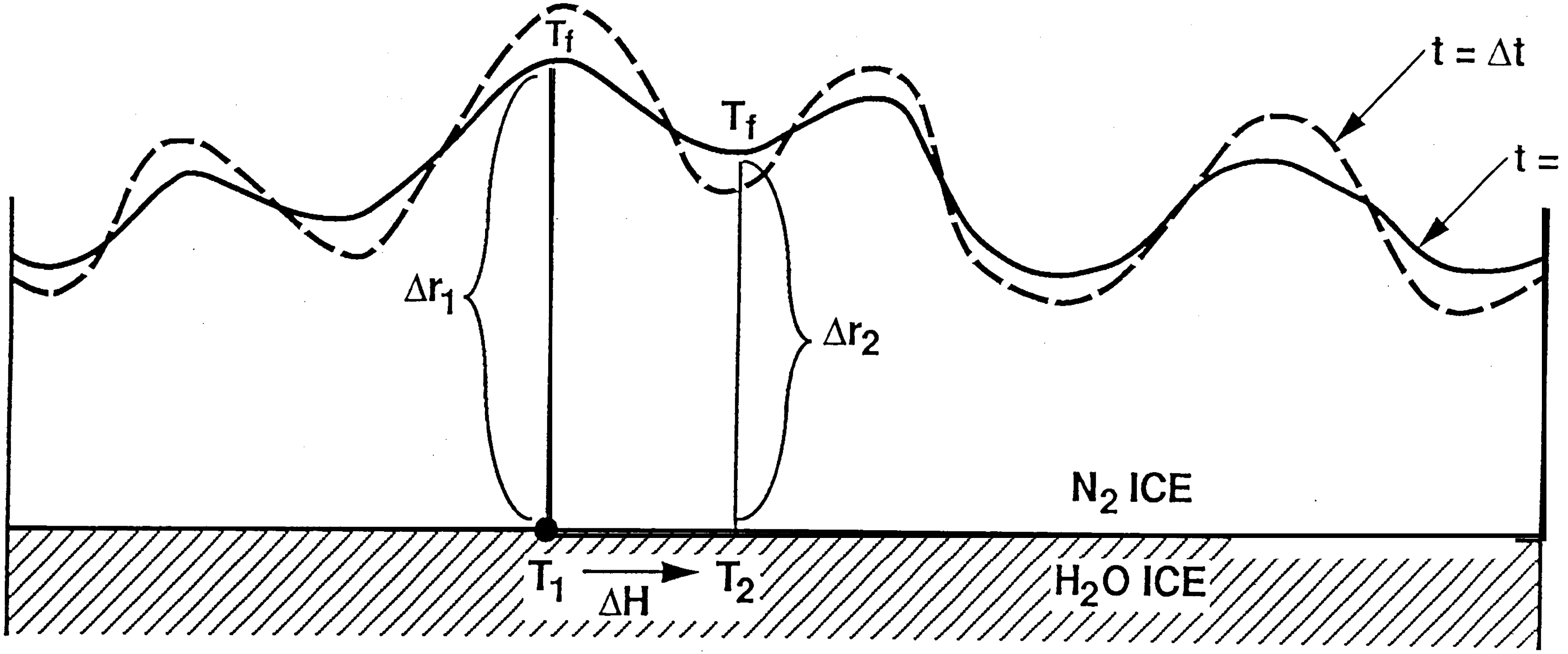
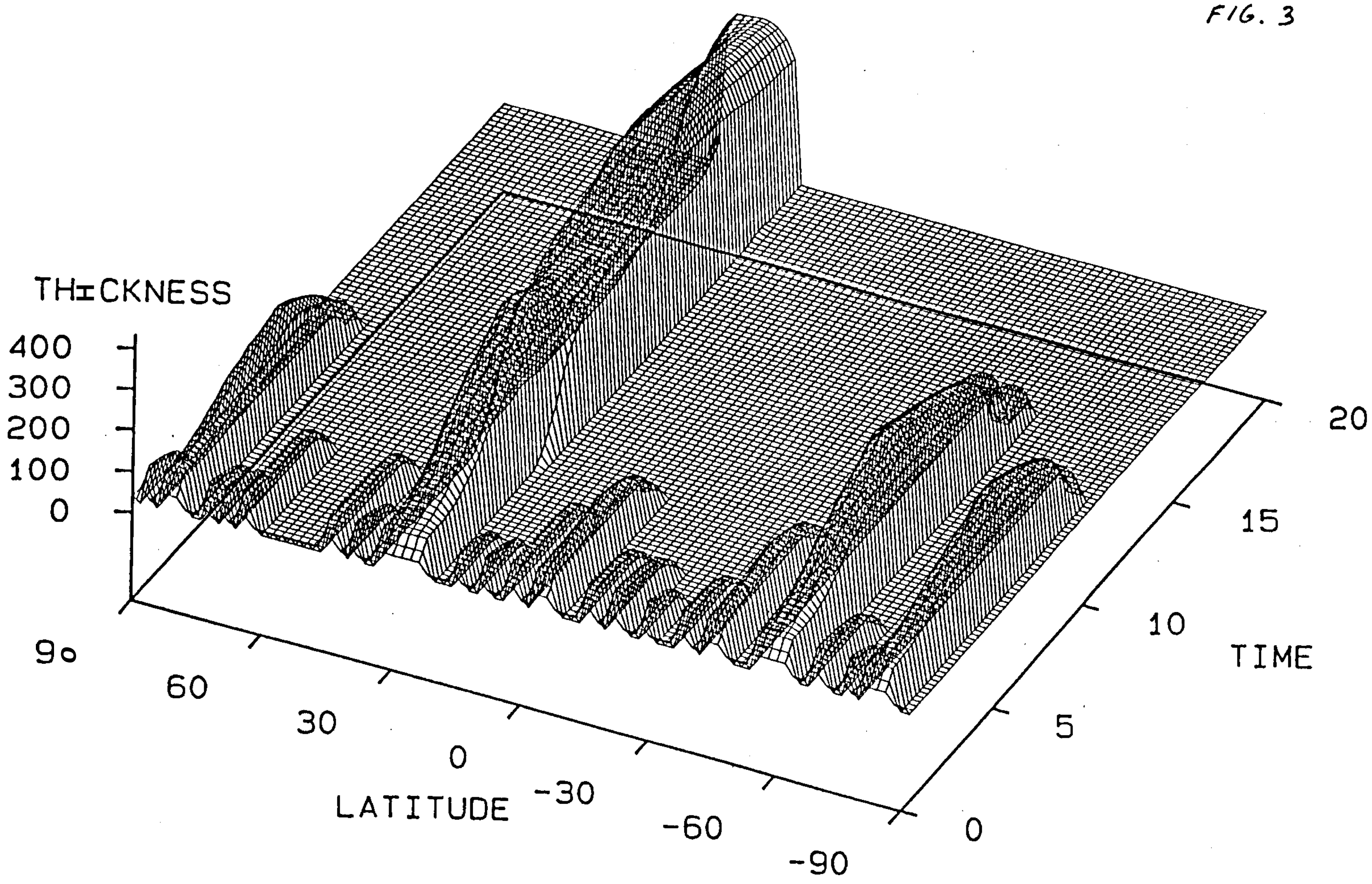

MANTLE
($H_2O$)
CORE
(SILICATES)
$N_2$ ICE

FIG -.2

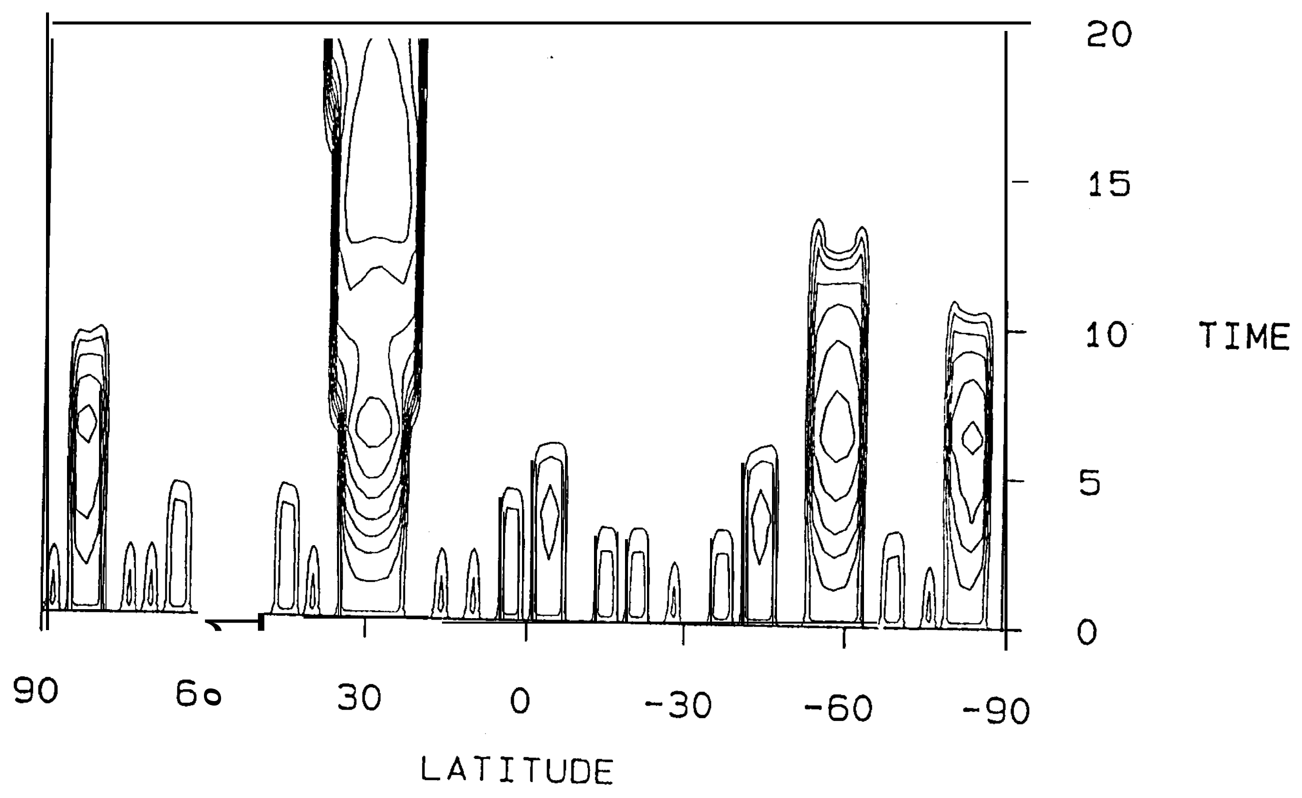



$\kappa \frac{(T_1 - T_f)}{\Delta r_1} = H = \kappa \frac{(T_2 - T_f)}{\Delta r_2}$ $\Rightarrow$ FOR CONSTANT k, IF $\Delta r_1 > \Delta r_2$ THEN $T_1 > T_2$, INDUCING LATERAL HEAT FLOW FROM THICKER AREAS TO THINNER AREAS. THUS HEAT FLOW CAN'T BE ISOTROPIC AND SYSTEM IS UNSTABLE $\Rightarrow$ THICKER AREAS GAIN MATERIA AT THE EXPENSE OF THINNER AREAS, EVENTUALLY COALESCING INTO ONE SPOT

FIG. 3



THICKNESS
400
300
200
100
0
90
60
30
0
-30
-60
-90
LATITUDE
0
5
10
15
20
TIME

20
15
10
5
0
TIME
90
60
30
0
-30
-60
-90
LATITUDE

FIG. 4

TEMPERATURE (K)

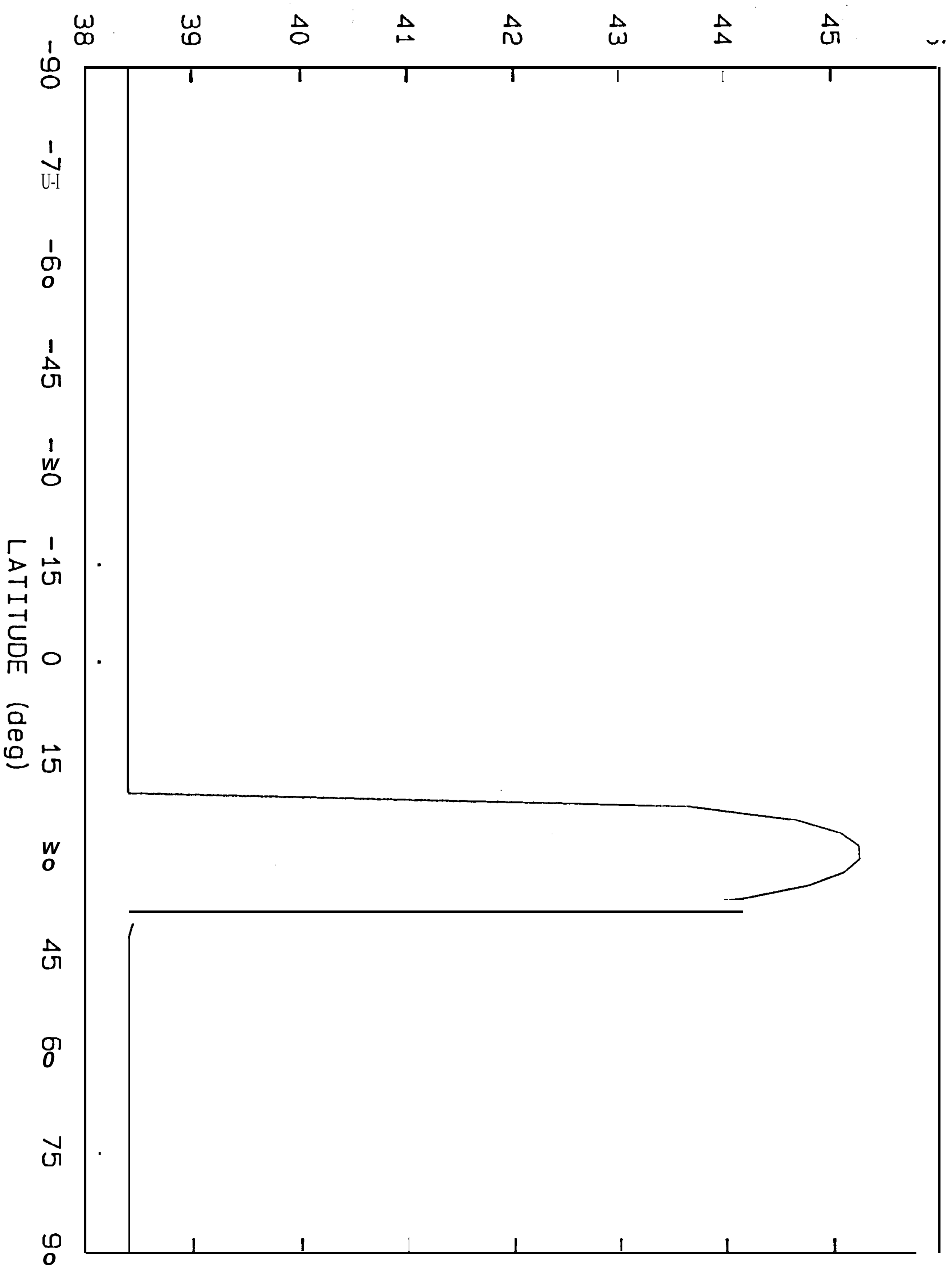
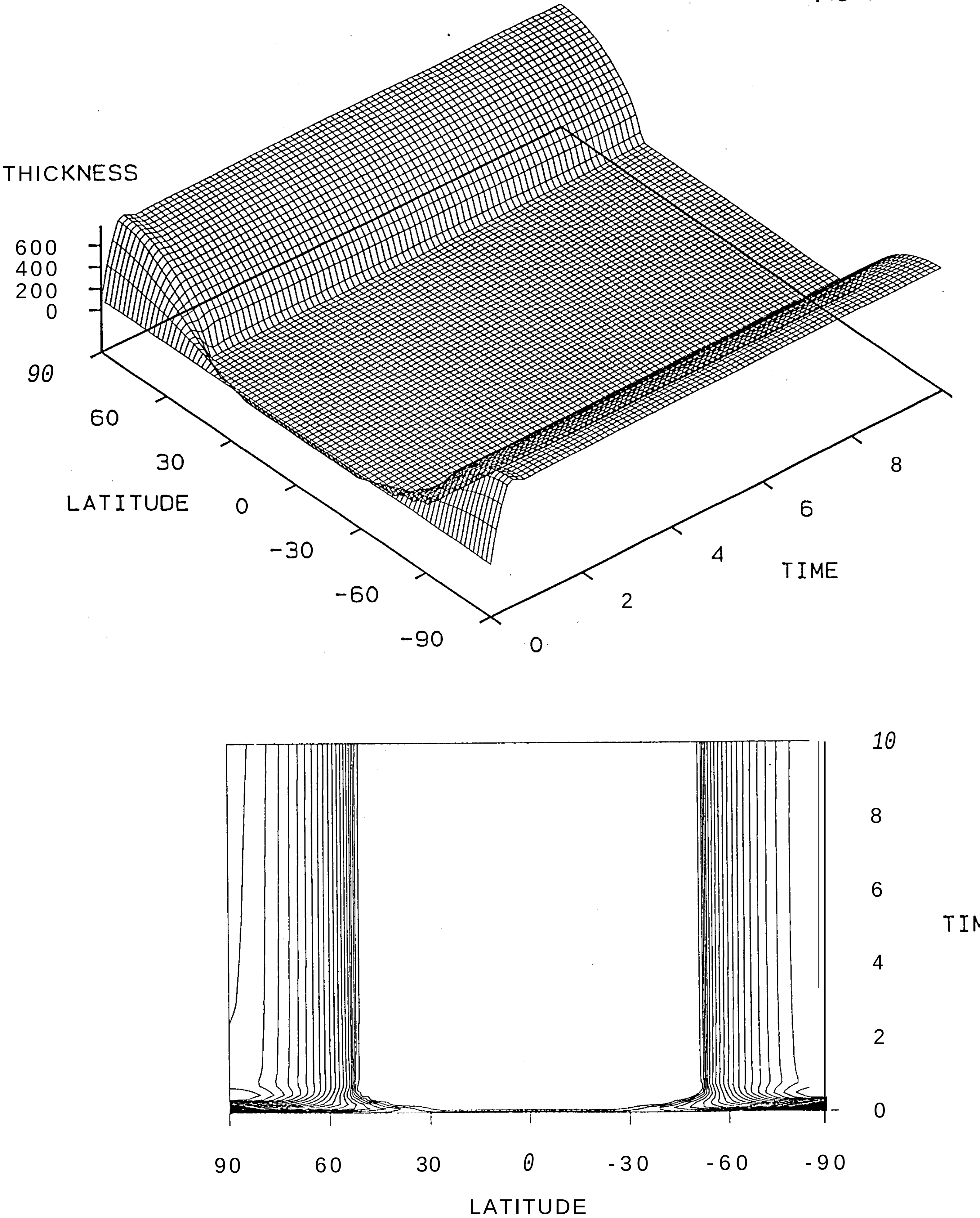


FIG. 5



THICKNESS
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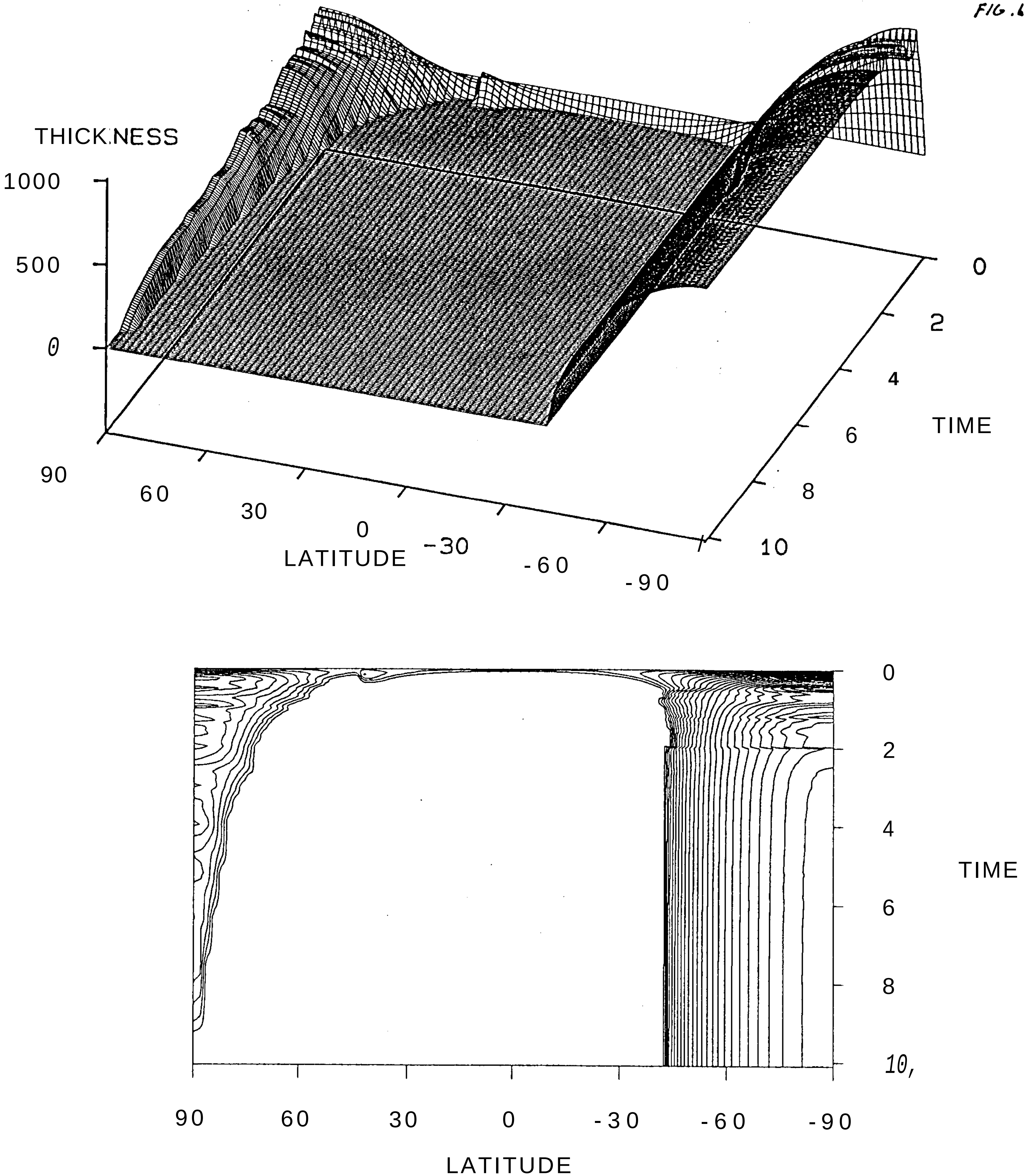
400
200
0
90
60
30
LATITUDE
0
-30
-60
-90
0
2
4
6
8
TIME
10
8
6
4
2
0
90
60
30
0
-30
-60
-90
LATITUDE

FIG. 6



THICKNESS
1000
500
0
TIME
0
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8
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LATITUDE
90
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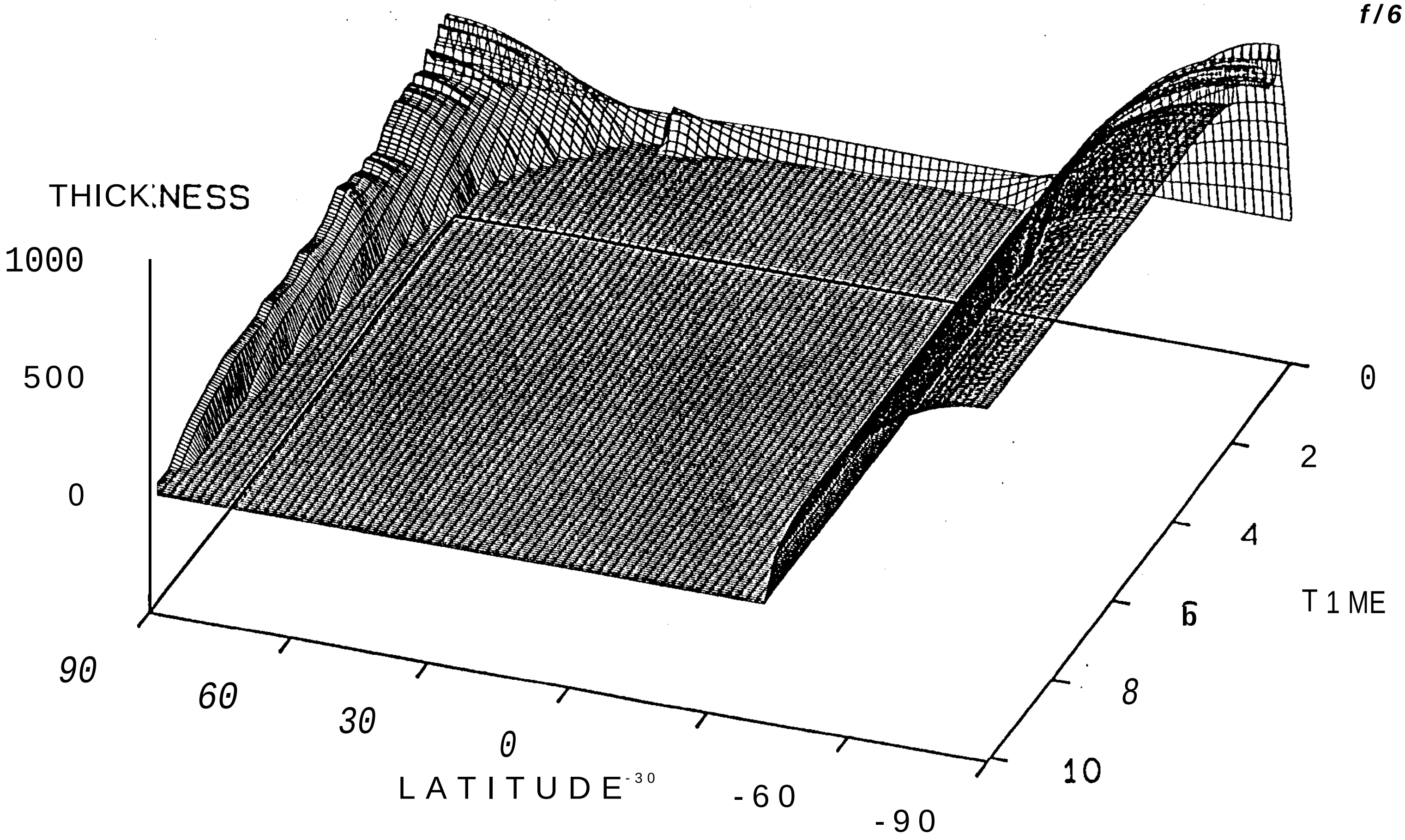
-90
LATITUDE


THICKNESS
1000
500
0
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2
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TIME
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LATITUDE
-30
-60
-90

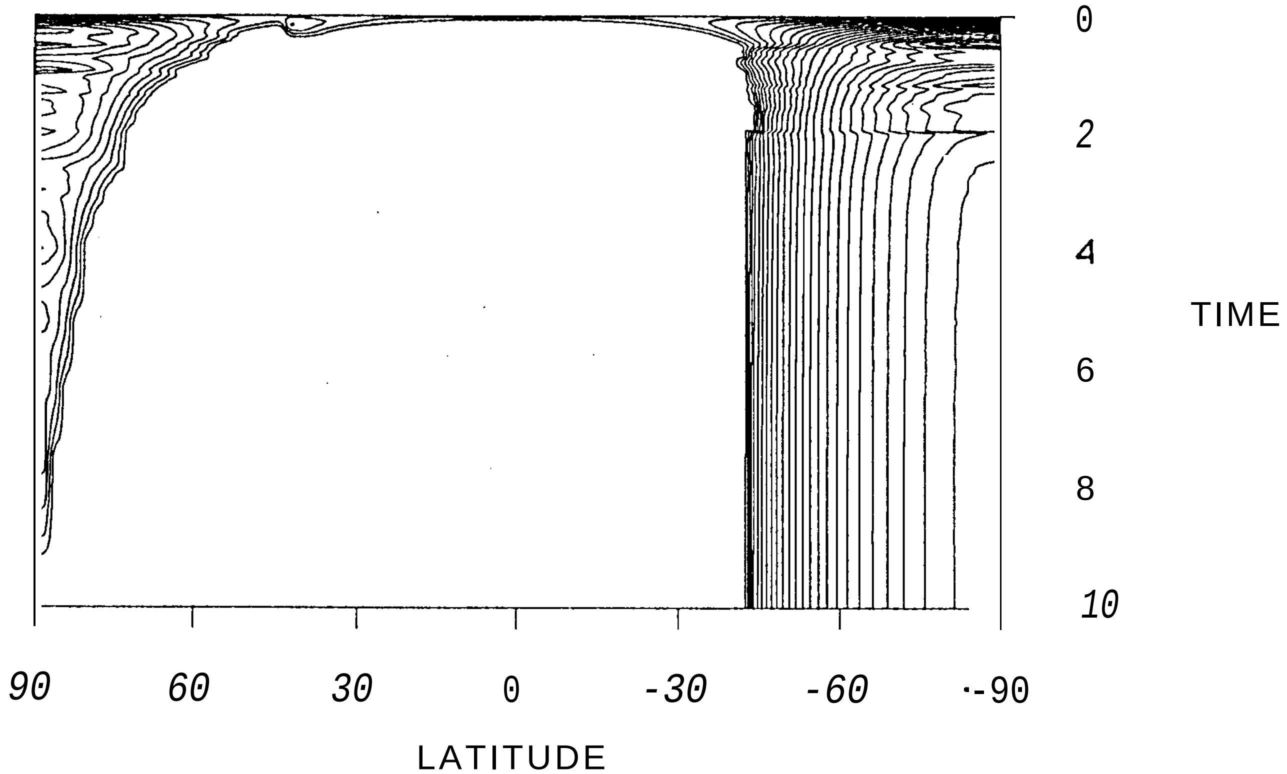
0
2
4
TIME
6
8
10
90
60
30
0
-30
-60
-90
LATITUDE

FIG. 7

HEATFLOW